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Journal

Journal of Geophysical Research: Biogeosciences, 124(1)

ISSN

2169-8953

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Publication Date

2019

DOI

10.1029/2018JG004566

Peer reviewed

RESEARCH ARTICLE

10.1029/2018JG004566

Special Section:

Biogeochemistry of Natural
Organic Matter

Key Points:

- Most carbon and nitrogen eroded from the upland catchments was either unprotected or physically protected inside weak macroaggregates
- Amount of eroded organic matter varied by year, but stabilization mechanisms did not vary consistently with type or amount of precipitation
- Organic matter in sediment was more readily decomposable than soil organic matter under similar environmental conditions

Supporting Information:

- Supporting Information S1

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Citation:

Stacy, E. M., Asefaw Berhe, A., Hunsaker, C. T., Johnson, D. W., Meding, S. M., & Hart, S. C. (2019). Stabilization mechanisms and decomposition potential of eroded soil organic matter pools in temperate forests of the Sierra Nevada, California. *Journal of Geophysical Research: Biogeosciences*, 124, 2–17. <https://doi.org/10.1029/2018JG004566>

Received 26 APR 2018

Accepted 31 OCT 2018

Accepted article online 20 NOV 2018

Published online 3 JAN 2019

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(continued)

Stabilization Mechanisms and Decomposition Potential of Eroded Soil Organic Matter Pools in Temperate Forests of the Sierra Nevada, California

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Abstract The lateral destination and potential decomposition of soil organic matter mobilized by soil erosion depends on factors such as the amount and type of precipitation, topography, the nature of mobilized organic matter (OM), potential mixing with mineral particles, and the stabilization mechanisms of the soil OM. This study examined how the relative distribution of carbon (C) and nitrogen (N) in different OM fractions varied in soils from eroding slopes and in eroded sediments in a series of low-order forested catchments in the western Sierra Nevada, California. We found that precipitation amount played a major role in mobilizing OM. More than 40% of the OM exported from these forested catchments was free particulate OM, or OM physically protected inside relatively less stable macroaggregates, compared to OM inside microaggregates or chemically associated with soil minerals. Years with high amounts of precipitation generally transported more mineral-associated OM, with lower C and N concentrations, while sediment transported in drier years was more enriched in unprotected, coarse particulate OM derived from surficial soils. When incubated under the same conditions, sediment C (from material captured in settling basins) produced 72–97% more CO₂ during decomposition than soil C did. Our results suggest that without stabilization through burial or reconfigured organomineral associations, this sediment OM is prone to decomposition, which may contribute to little to no terrestrial CO₂ sink induced from erosion in these Mediterranean montane forest ecosystems.

1. Introduction

Recent work on the role of erosion in biogeochemical cycling has been partly motivated by the potential for soil erosion to induce a terrestrial sink for atmospheric carbon dioxide (CO₂; Berhe et al., 2007; Harden et al., 1999, 2008; van Oost et al., 2007; Z. Wang et al., 2017). Soil erosion can lead to net removal of CO₂ from the atmosphere if some of the laterally transported soil organic matter (OM) is protected in slower cycling pools at the site of deposition and is replaced by the production of new photosynthate (Berhe et al., 2007). The potential for erosion to induce a terrestrial CO₂ sink has been demonstrated in both agricultural lands and undisturbed grasslands (Berhe et al., 2007, 2008; Harden et al., 1999; Nadeu et al., 2012; van Oost et al., 2007, 2012). Previous studies have also shown that the process of soil erosion can change the stability and stabilization mechanisms of OM (Berhe et al., 2012). Erosion can weaken physical stabilization of soil carbon (C) by aggregate disruption during raindrop impact or transport. In contrast, erosion may stabilize OM through burial or through association of C with reactive minerals at lower-lying depositional landform positions (Berhe & Kleber, 2013). Erosion can also affect growth-limiting elements such as nitrogen (N) eroded with topsoil, influencing soil N distribution across landform positions and among soil fractions and thus altering the rate of N turnover in soil (Berhe et al., 2018; Berhe & Torn, 2017). Mean basin slope, as a proxy for soil erosion, is a significant predictor of aboveground biomass in the Sierra Nevada (Milodowski et al., 2015) suggesting that erosion and the redistribution of plant-essential nutrients like N should be evaluated in a landscape-scale assessment of ecosystem productivity (Berhe et al., 2018; Berhe & Kleber, 2013; Berhe & Torn, 2017).

The ultimate effect of soil erosion on the fate of OM in eroding landscapes depends on many characteristics of soil, topography, landscape, and climate. The amount of OM dislodged and transported depends on the

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propensity of soil to erode, the vegetative cover above the soil, and the precipitation amount and intensity (de Vente et al., 2013; Meade et al., 1990; Pandey et al., 2016). Meanwhile, the tendency toward decomposition or stabilization of eroded OM depends on the nature of OM eroded, precipitation amount and intensity, soil texture and mineralogy, the degree of particle or aggregate detachment, the distance and mode of transport, and nature of the depositional landforms where eroded topsoil is deposited (Berhe et al., 2018). Preferential transport, deposition, and modification during and after transport alters eroded OM. Small and light organic and mineral particles remain suspended longer and can therefore travel farther than heavier particles (Verstraeten & Poesen, 2000). Transport of fine organic particles further downstream or decomposition during transport contributes to sediment C concentrations that are lower than the contributing soil source areas. Conversely, preferential deposition of heavier mineral fractions behind small in-channel dams can result in C enrichment in exported material (Nadeu et al., 2011, 2012). The erosivity of rainfall interacts with local topography, soil cover, and soil erodibility, to affect rate of soil erosion, nature of eroded material, and the ultimate depositional setting for eroded topsoil material. Highly erosive rainfall in steep slopes is likely to mobilize more material and move it farther than low erosivity rainfall in gentler sloping hillslopes. Furthermore, the lateral transport of topsoil and associated OM changes the microclimatic conditions that OM experiences. Erosion can move OM from typically well-drained eroding hillslopes to lower-lying depositional landform positions, which are more likely to experience flooding or anoxic conditions (Berhe & Kleber, 2013). Material deposited in midslope hollows showed less stabilization through physical and chemical mechanisms than soil in the toeslope plain (Berhe et al., 2012). Once deposited, the association of this eroded OM with soil minerals can lead to stabilization of OM through formation of new or reconfigured mineral-OM associations (Berhe et al., 2012). These physical or chemical associations with minerals can then limit the accessibility of eroded OM to further microbial decomposition (Grandy & Neff, 2008; Kleber, 2010; Kleber et al., 2011; Kögel-Knabner et al., 2008; Rumpel et al., 2004; Schmidt et al., 2011; Sollins et al., 2006, 2009).

Soil particle size distribution (PSD; i.e., soil texture) plays a key role in determining the effectiveness of different OM stabilization mechanisms. Studying dynamic landscapes in temperate grasslands, Berhe et al. (2012) found that long OM residence times were associated with aggregation and cation bridging in eroding positions with approximately 270 g/kg clay. By contrast, burial and association of OM with metal oxides were dominant OM stabilization mechanisms in low-lying depositional areas, averaging 430 g/kg clay (Berhe et al., 2012). Soil texture can also play a key role in protecting OM within aggregates stable enough to withstand detachment and erosional transport, as shown in a semiarid Mediterranean catchment in Spain with 120–137.5 g/kg clay (Nadeu et al., 2011). Reactive minerals, especially Fe oxides, can provide sorptive surfaces for stabilization of OM. For example, in temperate forest soils in Germany (44–597 g/kg clay), OM stabilization was facilitated by association with polyvalent cations and poorly crystalline metal oxides (Kaiser et al., 2011; Kayler et al., 2011). Despite the importance of soil texture on the nature and rate of soil erosion processes, most of the available data on stability and stabilization mechanisms of OM in eroding landscapes is from ecosystems with relatively fine-textured soils (e.g., Berhe et al., 2008, 2012; Harden et al., 1999; most sites studied by Kaiser et al., 2011). Comparatively few studies have assessed the effects of erosion on OM dynamics in coarse-textured soils (Lybrand et al., 2017; X. Wang et al., 2018), even though fine clay particles play an important role in the persistence of OM. Minimal clay and oxide concentrations in these soils could limit the potential for OM stabilization through surface bonding or association with metal oxides or through aggregate formation. The granite-derived soils of regions like the Sierra Nevada, for instance, have only 40–80 g/kg clay (Dahlgren et al., 1997; Jenny, 1941; Rasmussen et al., 2006). While erosion in small catchments in the southern Sierra Nevada has been previously studied (Hunsaker & Neary, 2012; Martin et al., 2014; Martin & Conklin, 2018; Stafford, 2011), it is not clear how coarse soil texture would impact OM stabilization through the erosion process. In addition to the lower clay concentrations, considerable lateral redistribution of C by soil erosion may occur because of the high C concentrations in the surficial soil horizons and relatively steep slopes that are prevalent in these landscapes (McCorkle et al., 2016; Stacy et al., 2015) and intense fall or winter precipitation events that may erode more material (Eagan et al., 2007), resulting in C and OM decomposing if not stabilized downslope.

The objective of this study was to determine how erosion transports different pools of soil OM in forested montane ecosystems with coarse-textured soils. Sequential OM extraction yielded fractions according to the degree of stabilization in the environment: free, physical protection within aggregates or chemically associated with the mineral fraction. We analyzed sediments collected from in-stream settling basins at

the outlets of eight low-order forested catchments in the Sierra Nevada where the soils were predominantly clay-poor Inceptisols derived from granitic parent material. The in-stream basins permit settling of material that reaches the streams and would otherwise be transported out of the catchments to another point downstream without the settling basins. Previous analysis of the sediment bulk composition showed that sediment export varied logarithmically with discharge and that sediment export and discharge were inversely related to sediment OM concentrations and fine mineral particles (Stacy et al., 2015). Concentrations of C and N in the sediment ranged from 15 to 190 g C/kg and 0.5 to 7.1 g N/kg, respectively. Based on radiocarbon and stable isotopes, McCorkle et al. (2016) identified the forest organic horizon and stream bed and banks rather than the surface mineral soil as the primary contributors to sediment. We expand on these previous works with a sequential separation of sediments (transported in 2009 and 2010) and soils C and N fractions based on ecologically relevant stabilization mechanisms, such as solubility, aggregation, and mineral association. In this work, we aim to determine the following: (1) how the relative importance of OM stabilization mechanisms (specifically for C and N fractions) in eroded sediments changes from year to year as a proxy for differences in precipitation; (2) how the relative significance of these OM stabilization mechanisms varies across the landscape, particularly across the rain-snow transition zone; (3) how the relative prominence of these OM stabilization mechanisms differs between eroding hillslope soil profiles and eroded sediments; and (4) how the decomposability of OM varies between soil (before erosion) and sediment (collected after erosion). We expected sediment OM stabilization mechanisms to vary year to year with annual discharge and across elevations based on the partition between rain and snow (vegetation differences were mitigated between elevations through site selection—see section 2.1). We predicted most OM would be in free organic particulate form due to the previously established importance of organic horizons compared to mineral soil. Mineral-associated OM transported with erosion was expected to increase in higher discharge years and increase as the rain-to-snow ratio increases, because more and coarser mineral-associated OM could be entrained during erosion, while fine minerals, and free, organic particles represent a smaller fraction of the total sediment during these times (Bracken et al., 2015; Rose, Karwan, & Aufdenkampe, 2018). Coarse soil texture would constrain the formation of microaggregates (*MiA*) and organomineral bonds, but we still anticipated that after erosional disturbance these fractions would be more important than less stable macroaggregates (*MaA*) for OM protection. Mineral-protected OM is expected to be less likely to decompose than free OM, and the process of erosion could disrupt aggregates and expose material buried at depth to expose previously protected material (Berhe & Kleber, 2013). Hence, we expected sediments to be more decomposable than OM in soils, and OM contained in those sediments in high-flow years (with more mineral-associated OM) to be less decomposable than sediments in low-flow years.

2. Materials and Methods

2.1. Site Description

This study was conducted at the Kings River Experimental Watersheds (KREW), which are funded by the U.S. Forest Service Pacific Southwest Research Station (www.fs.fed.us/psw/topics/water/kingsriver). Our study sites are within the Sierra National Forest (37.012°N, 119.117°W, Figure 1). The Southern Sierra Critical Zone Observatory (www.criticalzone.org/sierra) overlaps with the lower-elevation Providence catchments. The eight forested catchments (48–228 ha) in this study are grouped at two elevations across the rain-snow transition zone.

Precipitation falls in a Mediterranean-type climatic pattern, with little precipitation during the summer growing season. Between 75% and 90% of precipitation falls as snow at the higher-elevation Bull Creek catchments (2100–2490 m), while snow accounts for 20–50% of precipitation at the lower-elevation Providence Creek catchments (1,490–2,100 m; Hunsaker et al., 2012). Annual precipitation is quantified by Water Year (WY) beginning October 1 of the previous year. From WY 2004–2016, mean annual precipitation was 127 cm at the lower-elevation catchments and 138.6 cm at the higher-elevation catchments (Safeeq & Hunsaker, 2016). All catchments were relatively undisturbed during this study period, with the only significant source of human disturbance at these sites being U.S. Forest Service roads. Vegetation in all catchments is predominantly Sierra mixed conifer, with other coniferous forest types; vegetation differences were minimized through site selection during the establishment of the KREW project (Hunsaker, 2007). Entisols and Inceptisols in the Shaver (coarse-loamy, mixed mesic Pachic Xerumbrepts), Cagwin (mixed, frigid Dystric

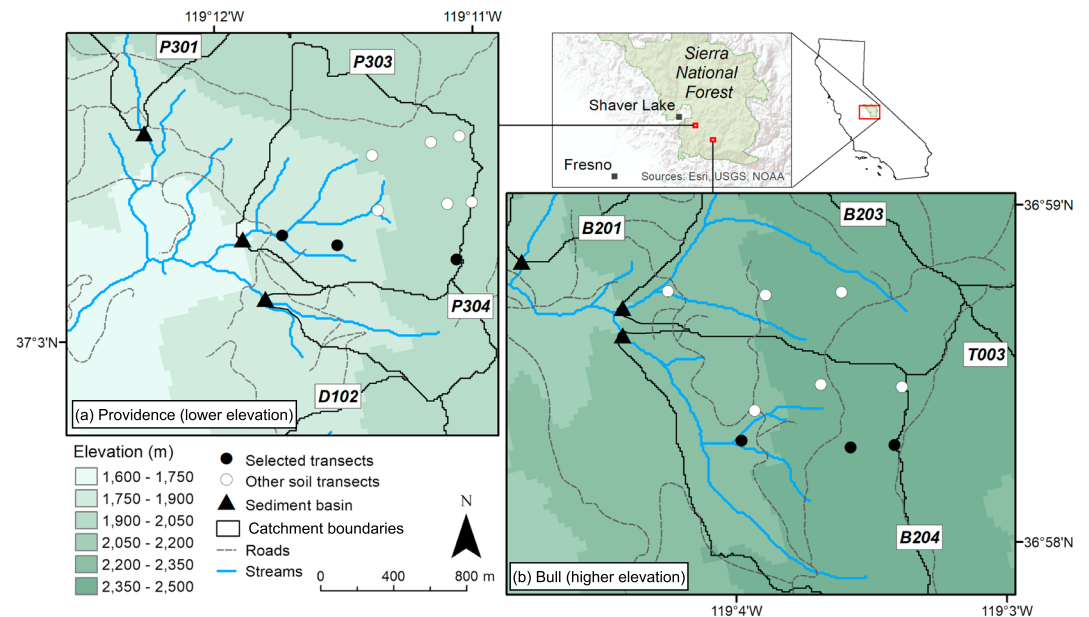


Figure 1. Map of study sites in Sierra National Forest, California, using a 30-m digital elevation model. Bulk composition of soil (circles) sampled at crest, backslope, and depositional landform positions was discussed in Stacy et al. (2015). A subset of these sites (filled circles) was sequentially extracted for comparison with sediments in this paper. This figure shows a portion of the total study site to better show the soil sampling areas; refer to Stacy et al. (2015) for a map of the full study site.

Xeropsammments), and Gerle (coarse-loamy, mixed, frigid Typic Xerumbrepts) series overlie Dinkey Creek Granodiorite. Johnson et al. (2011) provide a comprehensive description of the soil nutrient composition of these catchments.

To determine if erosion selectively transported one or more pools of C, we sequentially separated OM in soils from eroding slopes and sediments captured in settling basins. Sediment samples from water years 2009 and 2010 represent a contrast between an average precipitation year (WY 2009; 105 to 108 cm) and above-average precipitation year (WY 2010, 160 to 184 cm; Safeeq & Hunsaker, 2016). Sediment from WY 2011, used in the incubation but not the sequential extraction, had record-high precipitation (202 to 228 cm; Safeeq & Hunsaker, 2016). Sediment samples were collected from in-stream settling basins at all eight catchments to quantify the range of variation in the sediments across the landscape. Collection occurred in August or September, when the sediment basins were emptied for annual sediment load quantification; approximately 2 kg was representatively subsampled from the total material and returned to laboratory cold storage (4 °C), then subsequently sieved to 2 mm and air dried. Soil samples were sampled along hillslope transects in each catchment that were delineated based on location of comparable slope angle, aspect, and forest vegetation in each catchment. Soil samples were collected from points in 2011 representing crest (convex and eroding), backslope (sloped and eroding), and foot/toe-slope (concave and depositional) landform positions. The sampling location was selected at random within the landform positions, which varied in size from 50 to 2,000 m². Because of the time-intensive nature of the sequential extraction process, a subset of soil samples (three at each elevation) was selected for analysis from catchments B204 and P303 in KREW (hereafter referred to as the higher and lower catchments, respectively; identified in Figure 1). These two specific catchments were selected based on prior work that showed these catchments had almost average sediment production for their respective elevation catchment groups. The mineral soil was sampled in a single core by depth (0–10 and 10–20 cm) using a 10-cm diameter closed-bucket auger. Soils were sieved through 2-mm mesh and air dried. Also, due to the time-intensive nature of the fractionation, we processed sediment material from material already collected (WY 2009 and 2010; 16 samples), and later collected soils for comparison in 2011 (12 samples from six soil cores). We do not expect soil protection mechanisms to vary substantially across years but instead expect that variation in the relative importance of soil protection mechanisms would largely be due to differences in the amount and type of precipitation and ensuing erosion. More details

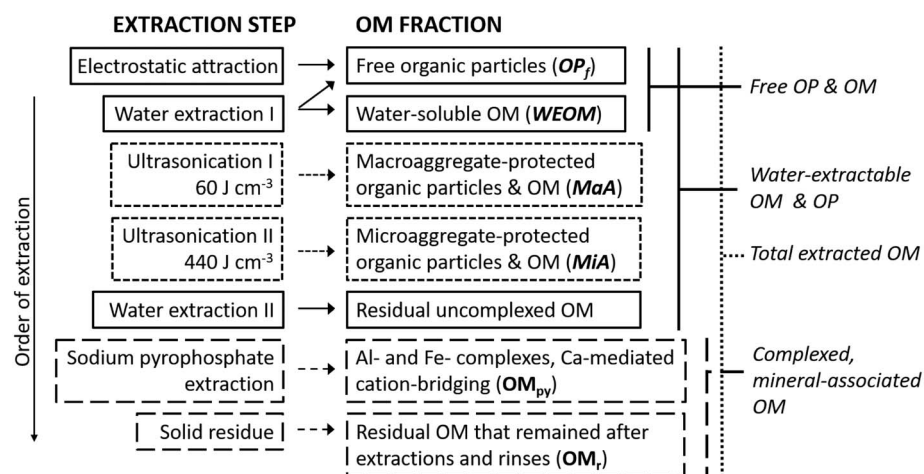


Figure 2. Diagram of the physical and chemical sequential separation steps and the OM fractions that were extracted from each step. OM = organic matter; OP = organic particles; OP_f = free organic particles; WEOM = water-extractable OM; MaA = macroaggregates; MiA = microaggregates; OM_r = residual organic matter.

regarding the sites, sampling methods, and soil and sediment (WY 2005–2011) properties can be found in Stacy et al. (2015).

2.2. Bulk Analysis

Sample C and N concentrations were measured using dry combustion (Costech ECS 4010 CHNSO Analyzer; Valencia, CA, USA) on oven-dried material, hand-ground with a mortar and pestle (Berhe et al., 2012). Sample pH was measured in a 2:1 water volume/sample weight slurry. The PSD was determined through laser diffraction (Beckman Coulter LS 13 320; Brea, CA, USA), and the specific surface area (SSA) was measured through N_2 adsorption (Beckman Coulter SA 3100; Brea, CA, USA).

2.3. Fractionation Scheme of C and N Analyses

Sequential physical and chemical separation techniques were used to isolate the C and N fractions that were unprotected, physically protected (through encapsulation inside aggregates), and chemically protected (by formation of bonds with soil minerals; Kaiser et al., 2009, 2010, 2011, and Kaiser, Ellerbrock, et al., 2012, Kaiser, Berhe, et al., 2012). From three 10-g analytical replicates, free organic particles (OP_f) were first separated by a combination of electrostatic attraction, density separation (water: 1 Mg/m^3), and sieving, after which free water-extractable OM (WEOM) was separated in a 6-hr extraction (extraction steps and fraction abbreviations in Figure 2). Aggregate-protected organic particles (OP) and OM were then separated by a combination of ultrasonication, density separation, sieving, and water extraction. Less stable MaA were first disrupted with sonication (60 MJ/m^3). A second round of sonication (440 MJ/m^3) released OM in more stable MiA. The OP and water-extractable OM in aggregate-occluded fractions are presented together as we presumed that the decomposability of aggregate-occluded OM depends first on aggregate breakup. Organic particles comprised 95% of C in all free OM, 96% of C in all MaA-protected OM, and 75% of C in the MiA-protected OM fraction (93%, 91%, and 52% of N, respectively). After the two sonication steps, a second water extraction step removed any other water-extractable material prior to mineral-bound OM extraction. Subsequently, samples were mixed with 0.1 M sodium pyrophosphate ($Na_4P_2O_7$) to extract OM associated with soil mineral surfaces through cation-mediated interactions (i.e., cation-bridging, crosslinking, and chelation; OM_{py}). The values reported here represent the combination of two 6-hr pyrophosphate extractions. Residual OM (OM_r) was defined as OM that persisted in the solid sample after sonication and water- and pyrophosphate-based extractions. Detailed descriptions of the operationally defined pools of soil OM, the different techniques used to separate them, and their ecological significance are given in Kaiser et al. (2009, 2010).

Separated OP fractions, OM_{py} solids, and OM_r were freeze-dried and weighed, then the total C and N concentrations of these fractions were quantified by dry combustion like the bulk soil. Each fractionated analytical replicate was run individually on the Elemental Analyzer, except where two of the three replicates had

individual weights less than 15 mg; this occurred 33% of the time, and in these cases all analytical replicates were combined.

Water- and Na-pyrophosphate-extractable OM fractions were filtered through Supor 450 filters (Gelman, 0.45 μm , hydrophilic polyethersulfone) using a vacuum pump. The C and N concentrations in the filtrates of *WEOM* and *OM_{py}* were measured using the nonpurgable organic C method on a TOC-VCSH Total Organic Carbon Analyzer interfaced with a TNM-1 unit (Shimadzu; Kyoto, Japan).

The mass recovery for all samples, including soluble OM and organic particles, was 91–103%. The overall recovery of bulk C was 60–99% (except for one sample with abnormally high recovery) and 59–119% for bulk N. The high variability in bulk C and N was due in part to the small size of some fractions in the initial sample replicate of 10-g material; the few samples with recovery >100% may indicate contamination. The C and N recovered from the second water extraction, a cleaning step, were excluded from statistical analysis in part because the fraction was small (less than 0.2% extracted C and 1.5% extracted N). The protection mechanism in this step was ambiguous, as it could have derived from the initial water extraction or one of the two subsequent aggregate dispersal steps, so material from this step cannot be assigned definitively to the water extraction or to aggregates disrupted through sonication. All values were expressed on an oven-dry (105 °C) mass basis.

2.4. Laboratory Incubation

Bulk soil and sediment samples were incubated in the laboratory to quantify the short-term potential for OM decomposition. We analyzed soil from the hillslope points used in the sequential extraction and soil from two additional hillslope transects at each elevation assessed in Stacy et al. (2015; Figure 1). We incubated sediment from WY 2009 (an average-flow year), 2010 (an above-average-flow year), and WY 2011 (an exceptionally high-flow year). Samples ($n = 54$) were incubated over a 43-day period under conditions of constant temperature (22 ± 1 °C) and water content (saturated). For each sample, deionized water was added to 10 g sieved, air-dried samples contained in a specimen cup (120 ml) until saturated (denoted by a slight glistening of the material surface). Samples were incubated under saturated conditions to reflect the state in which sediments generally occur in situ. Soils were also incubated at this water content to allow comparison of OM decomposition rates with those observed in sediments. Specimen cups were sealed within 0.97 L glass Mason jars with rubber septa in the lids. Thirty milliliters of deionized water was added to the bottom of the jars to help maintain the moisture content during the incubation. Headspace gas was sampled after 3, 6, 10, 17, 26, 35, and 43 days using a needle and gas-tight syringe. Carbon dioxide (CO_2) in the headspace samples was measured on a Shimadzu GC-2014 (Kyoto, Japan) gas chromatograph fitted with a thermal conductivity detector. After each sampling period, the jars were flushed with ambient air and then resealed. Cumulative CO_2 -C efflux over the incubation period was expressed relative to the total C concentrations of the material in to evaluate the decomposability of the OM (Grady & Hart, 2006). This cumulative CO_2 -C is hereafter referred to as mineralizable C.

2.5. Data Analysis and Statistics

Sediment data for elevation groups are presented by water year, with the values indicating field replicate means (\pm standard error; $n = 4$ catchments). A general linear model was used to establish the significance of elevation and year as explanatory factors for bulk sediment and the variation in C and N for each fraction independently. There was no interaction between year and elevation, so the interaction variable was excluded from the model. Where the model was significant, the Tukey-Kramer honestly significant difference test was used to differentiate means between elevation groups (high and low) and collection years (WY 2009 and WY 2010 for sequential extraction, all 3 years for incubation). For all statistical tests, an a priori α level of 0.05 was used to determine statistical significance. Statistical analyses were conducted using R 2.14.1 (www.r-project.org/). We expressed the C and N values of each fraction relative to the total C and N recovered from all fractions, rather than as absolute values, because we were interested in evaluating how erosion impacts the relative importance of stabilization mechanisms. Values reported for these fractions are analytical means ($n = 3$) and standard errors. Sample carbon-to-nitrogen (C:N) ratios are expressed on mass basis. Soil fractions from the 0–10 cm layer are compared to sediment fractions from the same catchments (values for 10–20 cm soil layer given in supporting information). No inferential statistics were calculated for soil and sediment comparisons because of the small number of samples analyzed. Results from the incubation are presented by

Table 1

Selected Soil (0–10 and 10–20 cm Mineral Soil Depths) and Sediment (Water Years 2009 and 2010) Characteristics.

Landscape position	Material	pH _w ^a	n	C ^b	N ^b	C/N ratio	Clay ^c	Silt ^c	Sand ^c	SSA ^c
				(g/kg)			(g/kg)			(m ² /kg)
<i>Higher elevation</i>										
Crest soil	0–10 cm	5.4	1	46.4	1.7	27.2	183	357	460	11,520
	10–20 cm	5.3	1	18.4	0.7	27.1	184	368	449	14,150
Backslope soil	0–10 cm	5.1	1	31.9	1.1	28.8	145	381	474	8,610
	10–20 cm	5.1	1	27.5	0.8	35.7	159	368	473	9,740
Depositional soil	0–10 cm	5.7	1	167.8	5.2	32.4	113	378	509	3,680
	10–20 cm	5.5	1	133.7	3.4	39.2	114	395	491	3,460
B204 sediment	WY 2009	5.0	1	64.3	2.5	25.3	58	233	709	1,990
B204 sediment	WY 2010	5.4	1	24.7	0.7	36.7	70	246	685	2,180
B201, B203, T003 sediments	WY 2009	5.0 (0.2)	3	72.5 (17.7)	3.2 (0.6)	22.9 (2.8)	87 (32)	294 (25)	620 (47)	2,993 (1,718)
B201, B203, T003 sediments	WY 2010	5.5 (0.1)	3	42.1 (18.5)	1.4 (0.5)	29.7 (3.9)	87 (18)	264 (33)	650 (43)	2,943 (1,065)
<i>Lower elevation</i>										
Crest soil	0–10 cm	6.2	1	54.0	2.6	20.9	117	365	518	6,960
	10–20 cm	5.4	1	35.6	1.6	21.6	106	371	523	9,270
Backslope soil	0–10 cm	6.3	1	85.4	3.4	25.1	122	375	503	11,990
	10–20 cm	6.5	1	32.2	1.4	22.8	106	371	524	17,090
Depositional soil	0–10 cm	6.5	1	9.7	0.5	19.2	163	378	459	7,120
	10–20 cm	6.1	1	33.3	1.2	26.8	162	374	464	9,170
P303 sediment	WY 2009	5	1	183.3	7.1	25.7	60	343	597	1,800
P303 sediment	WY 2010	5.5	1	125.3	4.1	30.7	83	331	587	2,270
P301, P304, D102 sediments	WY 2009	5.3 (0.2)	3	125.3 (9.7)	5.7 (0.9)	22.3 (0.4)	105 (25)	384 (1)	512 (24)	5,235 (1,931)
P301, P304, D102 sediments	WY 2010	5.7 (0.1)	3	60.4 (15.9)	1.9 (0.5)	31.3 (0.8)	86 (12)	281 (17)	634 (27)	3,017 (1,077)

Note. Soil landscape positions were selected as convex and eroding (crest), sloped and eroding (backslope), or concave toeslopes (depositional) along hillslope transects (Stacy et al., 2015). Values presented from the higher, snow-dominated catchment (B204) and the lower, rain-dominated catchment (P303). For comparison, sediment characteristics are also shown for the three other replicate catchments within each elevation group for the same water years (values represent group means (standard errors) across the three catchments). Values for <2 mm fraction. SSA = specific surface area.

^aStandard error ≤0.06 for analytical replicates ($n = 3$). ^bC and N concentration values represent mean of analyzed replicates ($n = 3$) conducted on a sample taken from each location (soils) or material from the collection basins (sediments). Standard errors ≤0.19 and ≤0.01 for C and N, respectively, for analytical replicates ($n = 3$). ^cBased on single analyses of field samples.

year and elevation for sediment, while the soil values are composited by elevation. No inferential statistics were computed for the soil incubation data because the soils were not sampled equally from all the catchments.

3. Results

3.1. Interannual Differences in Stabilization Mechanisms in Sediment

Sediment from the two years (WY 2009 and 2010) varied in both bulk physical and chemical characteristics but did not have as many statistically significant differences among fractions between the years. The sediment from WY 2009 had significantly higher total C and N concentrations than WY 2010 ($p = 0.014$ for C, and $p < 0.001$ for N) in the general linear model when elevation was also considered. PSD and SSA values for sediment, evaluated for the eight catchments, were different between years (Table 1). Year was a statistically significant predictor of the absolute amount of N in all sediment OM fractions except OP_f and the absolute amount of C in just two fractions, MaA and OM_r (C and N absolute concentrations in supporting information, Tables S1 and S2). As for the relative distribution of C and N among the fractions, there were no significant differences between the two years (Figure 2).

The WY 2009 sediment had significantly lower total C:N mass ratios than the WY 2010 sediment across all eight catchments ($p = 0.001$; Table 1). One catchment, B201, had a notably low C:N ratio in both years (17.4 in WY 2009 and 22.0 in WY 2010). Most catchments showed a similar pattern for C:N ratios in the separated fractions, except for the lower-elevation P303 and D102 catchments which had higher C:N ratios in WY 2009 than in WY 2010. Evaluated by elevation group, C:N ratios were lower in WY 2009 than in WY 2010 for five of six fractions; however, the difference was only significant for OP_f and MaA fractions

Table 2
 R^2 Values for the General Linear Model for Carbon (C) and Nitrogen (N), and the C:N Ratio for Bulk Samples

	C (g/kg)			N (g/kg)			C:N ratio		
	R^2	L-H	2009–2010	R^2	L-H	2009–2010	R^2	L-H	2009–2010
Total	0.524	*	*	0.656	*	*	0.504		*
OP_f	0.222	*		0.337	*		0.416		*
<i>WEOM</i>	0.216			0.744	*	*	0.287		*
<i>MaA</i>	0.586	*	*	0.586	*	*	0.241		
<i>MiA</i>	0.394	*		0.494	*	*	0.243	*	
OM_{py}	−0.051			0.522		*	0.025		
OM_r	0.702	*	*	0.670	*	*	0.048		
	C (% recovered)			N (% recovered)					
	R^2	L-H	2009–2010	R^2	L-H	2009–2010			
OP_f	−0.050	*	*	−0.089					
<i>WEOM</i>	−0.084	*		0.571		*			
<i>MaA</i>	0.255			0.347		*			
<i>MiA</i>	0.046	*	*	0.201					
OM_{py}	0.121	*		0.156					
OM_r	−0.022			−0.034					
Recovery rate	−0.121	*	*	0.094					

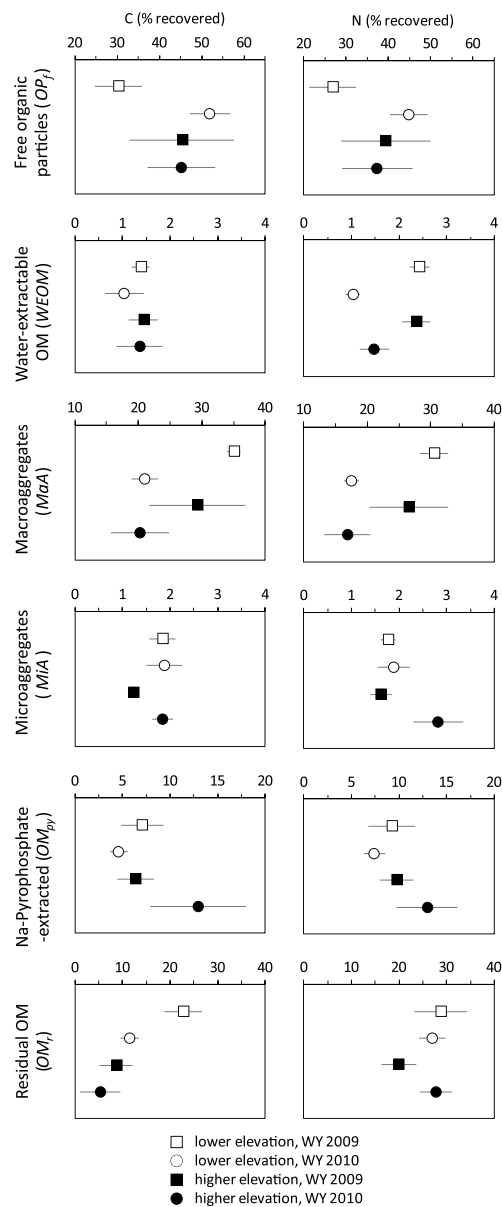


Figure 3. The distribution of carbon (C) and nitrogen (N) among extracted organic matter fractions for sediment collected from catchments in the lower-elevation and higher-elevation groups (symbols represent catchment means, and error bars denote ± 1 standard error, $n = 4$). Values are expressed relative to the amount of that element recovered from all fractions combined. Note variable axes.

elevation soil transect had higher C:N ratios than the lower-elevation transect across all depths and landscape positions.

Sediment PSD was predominantly sand (620 ± 18 g/kg), with comparatively little clay (83 ± 8 g/kg; Table 1). The only statistically significant difference between collection years for WY 2009, WY 2010, and WY 2011 sediment was the silt fraction between WY 2009 and WY 2011 ($p = 0.020$; WY 2011 data not shown). A difference between the elevations ($p = 0.017$) can be explained by much finer sediment at two catchments (P304 and B201) that skew the averages ($p = 0.046$ and 0.005 for P304 and B201, compared to the other catchments). Soils were also coarse but had higher amounts of clay (140 ± 8 g/kg) compared to the sediments (Table 1). With the caveat that the soils were only sampled in two of the eight catchments, the soils had finer texture and higher SSA than either sediment year ($p < 0.001$). The soil clay content varied by landscape position at both elevations. In the lower-elevation catchment, the clay content was lower in the crest and backslope positions than in the depositional position, but that pattern reversed at the higher elevation. These differences between landscape positions were not assessed statistically because there were so few replicates.

Sediment OM was largely located in free water-extractable material or in *MaA*. The organic particles from the *OP_f* and *MaA* fractions combined composed more than 66% of all sediment C (Figures 3 and 4). *OM_r* and *OM_{py}* were the next most important fractions. The *OP_f* and *MaA* fractions also comprised more than 55% of sediment N and, along with *OM_r*, were the three most important fractions for N. *WEOM* removed prior to aggregate disturbance was a small fraction and contained less than 2.5% of C and N.

Organic matter in the soils was mostly found in fractions with greater physical or chemical stabilization (Figure 4; 10–20 cm data in supporting information). In the soil samples, mineral-associated OM (*OM_{py}* and *OM_r*) accounted for 19% more of the recovered C and 22% more recovered N than in the corresponding sediment. The *OP_f* and *MaA* fractions, which were important for sediment C and N stabilization, accounted only for 45% of the C and 33% of the N in most soil samples (Figure 3; supporting information for 10- to 20-cm data). The higher-elevation depositional soil was most like the sediment, with a greater amount of C and N in the *OP_f* and *MaA* fractions. The C:N mass ratios in the soils and sediments from the lower-elevation catchment were generally similar, but the soils in the higher-elevation catchment had higher C:N ratios than the collected sediments.

3.4. Mineralizable C

Sediment had more mineralizable C ($35.3\text{--}73.9$ g $\text{CO}_2\text{-C/kg}$ soil C) than did soil ($21.9\text{--}39.2$ g $\text{CO}_2\text{-C/kg}$ soil C) over the course of the 43-day laboratory incubation (sediment $>0\text{--}10$ cm soil: $p = 0.01$; sediment $>10\text{--}20$ cm soil: $p = 0.001$; Figure 5). Mineralizable C in sediment was inversely related to annual stream discharge and showed a greater negative slope in the lower-elevation catchments (Figure 5). Sediment at lower elevations produced 11–42% more mineralizable C compared to the higher-elevation catchments in the same years. Like sediments, the soils from the lower-elevation catchments had 79% more mineralizable C for the 0–10 cm surface soils and 14% more at the 10–20 cm depth. However, these differences in mineralizable C between elevations were not statistically significant for either sediments or soils at either depth. Assessed independent of the soil values, sediment in WY 2009 had significantly more mineralizable C than WY 2011 ($p = 0.016$) when accounting for both elevation and year.

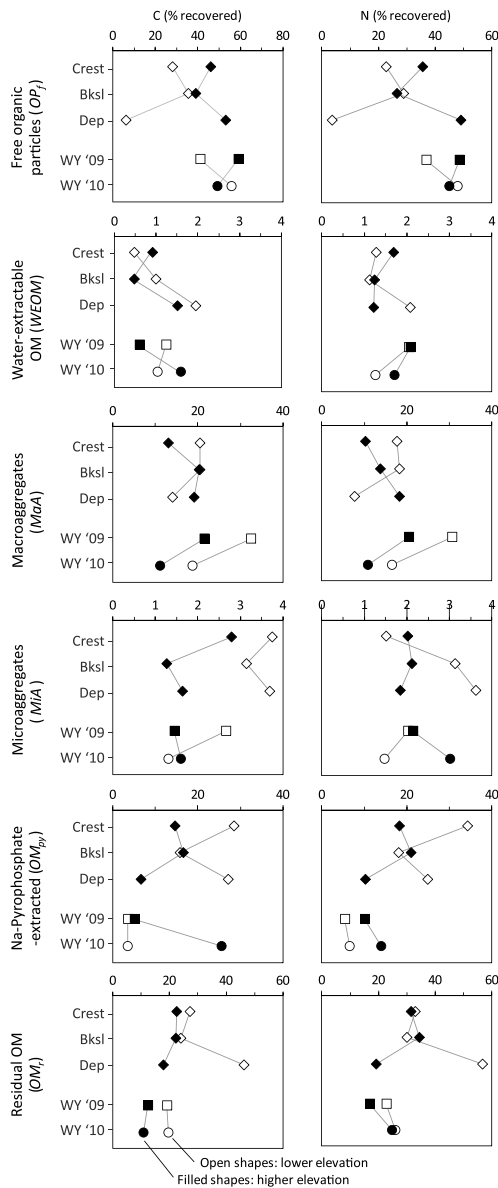


Figure 4. The distribution of carbon (C) and nitrogen (N) among extracted organic matter fractions for soil from three landform positions and sediment from water years 2009 and 2010 for the selected catchments P303 and B204. Values are expressed relative to the amount of that element recovered from all fractions combined. Note variable axes. (See section 2 and Stacy et al., 2015 for selection of crest, backslope, and depositional landform positions).

4. Discussion

4.1. Interannual Variability in Selective Transport of Different OM Fractions

We found that annual precipitation and discharge affect the total sediment OM concentration and the extent of different stabilization mechanisms but not the relative importance of those mechanisms. The average (WY 2009) and above-average precipitation years (WY 2010) yielded different amounts of OM but did not yield erosion of OM protected by one mechanism more than another. Annual sediment load for years 2005–2013 showed a positive correlation with precipitation, while the sediment production across all catchments varied inversely with C and N concentrations (Hunsaker & Neary, 2012; McCorkle et al., 2016; Stacy et al., 2015). We found that the relative concentrations of C and N in the OP_f , OM_{py} , and OM_r fractions reflect these same patterns.

The total sediment export and sediment composition are determined in large part by precipitation intensity and annual discharge (Boix-Fayos et al., 2009; Stacy et al., 2015). Important climatic variables that affect amount and nature of OM eroded include precipitation intensity, kinetic energy of falling rain, increased runoff during rain-on-snow events, and the snowmelt rate. At the higher-elevation catchment, daily precipitation exceeded 50 mm four times in WY 2009 and 10 times in WY 2010, while daily precipitation at the lower catchment exceeded that point three times in WY 2009 and seven times in WY 2010 (measured at the upper meteorological station for each catchment group; Hunsaker & Safeeq, 2018). In one notable storm in October, WY 2010, 171.7 mm fell in 1 day at the upper elevation catchment and 210.4 mm fell in the lower-elevation catchments. This high-intensity rainfall storm could have contributed to or increased the differences between the two years for sediment composition or transport from the catchment. At the same time, these values do not reveal other moderating factors, such as instantaneous precipitation intensity, the fraction of precipitation falling as rain compared to snow, or runoff rates. We acknowledge that erosion and sediment transport can vary widely between storms, and this would impact the OM composition and stabilization. Because sediment analyzed in the present study and previous studies (McCorkle et al., 2016; Stacy et al., 2015) was sampled by year, we evaluated only the amount of annual precipitation and the difference between elevations (as a proxy for precipitation form). In sediment (<2 mm) captured in these basins in WY 2005–2011, annual discharge was inversely related to bulk C and N concentrations, clay fractions, and SSAs (Stacy et al., 2015). Rose, Karwan, and Aufdenkampe (2018)

observed similar patterns with regard to discharge for suspended particulate matter over time in a single catchment in Pennsylvania, USA. They reported that larger storms resulted in higher contributions from upland sources, while low event flows produced suspended particulate matter that was finer and derived from in-stream sources. While year and elevation were significant predictors for total OM, they did not consistently predict differences in OM stabilization mechanisms. Our findings suggest that the mechanisms of OM stabilization in eroded sediments are dynamic and difficult to predict. A shift from snow to rain, and increasing storm intensity, may entrain larger particles and more mineral material, but this cannot be conclusively confirmed from our findings (see section 4.5 for a discussion of sediment basin efficiency and possible artifacts of the collection method). To resolve those differences, we recommend further resolution of stabilization mechanisms by sediment producing event, on suspended materials, or on materials captured in

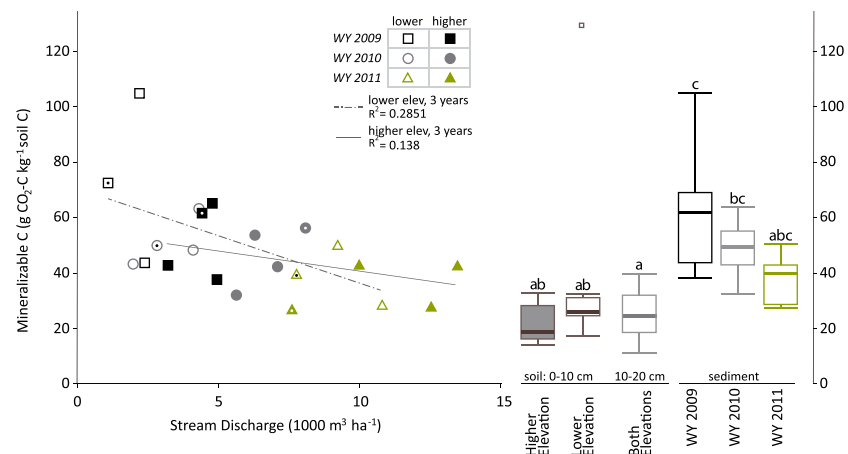


Figure 5. Mineralizable carbon in sediment declined with increasing stream discharge at both the higher- and lower-elevation catchments. Mineralizable C is expressed as cumulative carbon dioxide carbon evolved from the sediment and soil under saturated conditions during a 43-day laboratory incubation relative to total C concentration in the sediment or soil. Box plots are means and quartiles, with one outlier (10–20 cm values and sediments combined across elevations; letters mark significantly different means). Small dots mark the upper and lower catchments that were selected for the sequential extraction in 2009 and 2010.

transport across the uplands, in addition to sediment captured in the basins. Collection and assessment of the suspended materials would provide insight into what may be lost downstream without settling in the sediment basins. It would also be illuminating to test the separated fractions under controlled laboratory conditions through incubation or simulated transport experiments (exposure to aggregate disruption, mixing, and wetting/drying cycles) to isolate the fractions most prone to decomposition, and thereby assess the vulnerability to decomposition of the various stabilization mechanisms.

4.2. Effect of Precipitation Type on Nature of OM Mobilized by Soil Erosion

The location of the study catchments at high and low elevation (along the western Sierra Nevada climosequence) can be considered a surrogate for climate change, assuming higher-elevation catchments would behave more like lower-elevation catchments as the climate continues to warm (Hart & Perry, 1999). With warming, the dominance of snow at the higher-elevation catchments may transition to a mix of rain and snow, similar to that seen in the lower-elevation catchments, while precipitation in the lower-elevation catchments may fall entirely as rain. More rain or higher-intensity storms could increase erosion by altering soil vegetative cover and increasing erosion on hillslopes and in streams (Nearing et al., 2005; Pandey et al., 2016). However, while the total sediment transport increases with stream discharge (Hunsaker & Neary, 2012; Stacy et al., 2015), and the sediment C and N concentrations decrease with increasing discharge (Stacy et al., 2015), we did not see consistent patterns in the relative distribution of C and N among stabilization mechanisms for the two years studied here. Overall, our results showed that as with water year, elevation was more important as a factor for the bulk sediment composition than for the distribution of OM across stabilization mechanisms. McCorkle et al. (2016) assessed soil and sediment samples from catchments B203 (adjacent to the higher catchment selected here) and P303, as well as sediment samples from WY 2009–2012. In the higher-elevation site, the primary source of sediment was found to be surface (O) horizon material, while the lower-elevation site has a mix of contributing sources, including O horizon and stream bed and bank erosion (McCorkle et al., 2016). Our assessment of the WY 2009–2010 sediment similarly emphasized the relative importance of light and free OP—material that would largely come from O horizons—in catchments at both elevations over mineral-associated fractions that could come from deeper mineral horizons. The coarse mineral material we found in the sediment could be generated by stream bed or bank erosion rather than surface erosion of the O horizon or be the result of preferential settling forces in the sediment basin of the material that reaches the stream (discussed further in section 4.5). The stream bed in these forested headwater catchments likely offers only a temporary storage point for material eroded from uplands or banks (Martin & Conklin, 2018). In the lower-elevation catchments, Stafford (2011) found that erosion

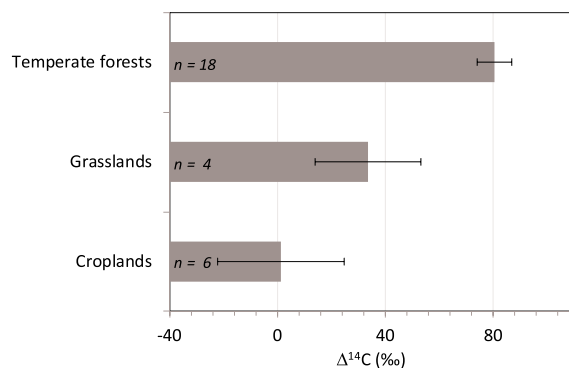


Figure 6. $\Delta^{14}\text{C}$ (mean \pm standard error) of eroded sediments or topsoil material (upper 10–20 cm) in eroding landform positions. The higher values for eroded sediments from forests suggest that the organic matter eroded from these ecosystems is of more recent origin than the organic matter eroding from grasslands or croplands. Data sources: Forest—McCorkle et al. (2016); grassland—Berhe et al. (2008, 2012); croplands—VandenBygaart et al. (2012, 2015).

features such as gravel and dirt access roads, another source of deep mineral material, were not connected to streams. While bulk materials differed with elevation, we found that elevation was not a determining factor for the relative importance of the stabilization mechanisms.

4.3. Selective Mobilization of Different OM Fractions With Soil Erosion

Based on previous studies, we expected the surface and stream-centric erosion processes at our study catchments to produce sediment enriched in easily decomposable light fractions and previously aggregate-protected OM. We did find that the light OP_f fraction was one of the largest fractions of sediment C and N, but transport of aggregate-protected OM was more complex. The soil OP_f at the lower-elevation catchment was comparable with the particulate C fraction observed in some other studies, which was as much as 14–40% of soil C (Boix-Fayos et al., 2009; Swanston et al., 2005) but more than the 2.9–3.75% C in the free light fraction reported by Berhe et al. (2012). Previous studies found preferential transport of light organic particles (equivalent to the OP_f fraction; e.g., Nadeu et al., 2012), as we did in this study, especially in the lower-elevation catchments. In the

erosion landform position soil profiles, the four largest OM fractions (OP_f , MaA , OM_{pyr} , and OM_r fractions) at the crest and backslope soils were largely similar to each other but different from the depositional soils. The relative importance of these fractions at the landscape positions reversed in the deeper soil layer (10–20 cm). This was not conclusive support of pre-erosion and posterosion patterns in stabilization mechanisms, as some of the soil values from depositional locations were more different from sediment than the eroding crest and backslope locations. We found no significant difference in the relative amount of C or N in the MaA and MiA across landform positions, and MiA fraction C and N are similar between sediments and soils. This finding suggests that these aggregates (MiA) are likely stable enough to withstand breakage during detachment and transport, even though they protect a relatively small proportion of the OM overall. Assuming most of the MiA fraction that was mobilized by erosion was largely captured in the sediment basins, this finding suggests that the coarse, less weathered soils in our study catchments have limited potential to physically protect C in MiA . Chemical organomineral associations can also play a role in erosional OM transport. Erosion represents one of the most important mechanisms by which otherwise stable, mineral-associated OM can be relocated in large quantity, and it could facilitate mineral-OM association by exposing new mineral surfaces for binding with OM, preferentially transporting oxides, and promoting chemical transformations during transport (e.g., Berhe et al., 2012; Kögel-Knabner et al., 2008). We found that complexed and residual C fractions were smaller in sediments than in the soils at the eroding landform positions, suggesting erosion, during years without major storm events and with no land disturbance, was not a major factor in mobilizing mineral-bound OM in coarse-textured soils common in montane forests of the temperate zone.

4.4. How Mineralizable Is the Eroded Organic Matter?

The mineralization potential of OM has implications for the net impact of erosion on the C balance in an ecosystem. Investigating the nature of OM eroded from temperate forested catchments in the southern part of the Sierra Nevada (McCorkle et al., 2016; Stacy et al., 2015; and present study), we found that most of the OM eroded from low-order catchments in the western slope of the Sierra Nevada is composed of relatively unprotected OM that matches O horizon material. The O horizon material has high concentrations of C and is free from chemical or physical association with soil minerals. Incubating this O horizon material, or dividing the surface soils into shallower layers, could more closely represent the material that is affected by erosion in any given year. The radiocarbon concentration of OM eroded from these forest ecosystems is much younger (modern, post-1950) than OM eroded from grassland and cropland systems in Northern America (Figure 6). If OM in upland forest systems is more susceptible to decomposition compared to that eroded from croplands or grasslands, it could mean that erosion is not likely to lead to a terrestrial C sequestration in such forested catchments. The turnover rate of soil OM can increase through detachment, aggregate breakdown, or a favorable change in microclimate conditions, or conversely can decrease through burial, aggregation, or

through chemical formation of stronger or new bonds with mineral surfaces (Berhe & Kleber, 2013). In the temperate forest catchments we studied, the eroded OM likely has low potential for posterosion stabilization because of the dominance of steep slopes, the small size of depositional areas near streams (minimizing the chance for physical stabilization), and the small SSA and clay fractions (minimizing the chance for chemical stabilization). Substantial decomposition during overland transport or residence in the stream is possible. High in-stream processing of coarse leaves and other particulate matter was seen in streams in Puerto Rico, resulting in lower sediment OM export rates despite high inputs of OM (Heartsill-Scalley et al., 2012). While the sediment basins do not elucidate how much OM decomposed during transit or under what conditions it will be deposited, results from our laboratory incubation showed that sediment OM was more likely to decompose than soil OM from the contributing slopes. The contribution of unprotected material from forest organic horizons (McCorkle et al., 2016), the large OP_f fractions in sediment, and the higher decomposability of the sediment in the incubation all indicate that there is limited potential for erosion to induce a net C sink in these temperate montane ecosystems.

4.5. Limitations of Methodology

Sediment basins located at the outlet of a catchment are useful for determining amount of sediment mobilized by erosion annually and assessing effect of changes in climate or land use. However, sediment transported to the basins is only sampled once at the end of the water year, and the sediment capture efficiency is typically less than 100% and variable over time. In addition, this study only evaluated a short period of natural variation. The KREW sediment basins were constructed in steep mountain streams, in compliance with constraints regarding environmental disturbance. These sediment basins have good capture efficiency for denser or larger mineral particles (bed load) but during high flows the capture efficiency is likely to be less than 100% for fine OM associated with suspended sediment and free particulate OM. During years with high stream velocity and discharge, fine minerals and low-density OP_f may not settle out of the water column within the basin, as they take longer to settle and settle at lower stream velocities and discharge (Haan et al., 1994). This is corroborated by the inverse relationship between stream discharge and bulk C and N concentrations, clay fractions, and SSAs (Stacy et al., 2015). Catch samples of suspended particulate matter during a range of stream conditions has demonstrated distinct sediment sources during high and low stream flows, as well as a loss of fine particulate matter during high stream flows (Rose, Karwan, & Aufdenkampe, 2018; Rose, Karwan, & Godsey, 2018). The interannual variation in sediment collection (in part representing settling basin performance or behavior) demonstrates a negative relationship with discharge (Stacy et al., 2015), but PSD was mostly statistically indistinguishable between the two years (2009 and 2010) evaluated in this study. By comparison, the differences between sediment and soils for PSD and SSA were much greater, with sediment being consistently coarser. If this pattern is an artifact of sediment basin capture efficacy, it is at least happening across all of the settling basins.

Sediment transport from these catchments may largely be episodic. A turbidity study assessing the effects of discharge in similar Sierra Nevada catchments found that turbidity occurred more with early-season fall and early/midwinter events (often rain) than during spring snowmelt (Martin et al., 2014). Turbidity was not measured in this study, and we cannot estimate what fraction of the eroded OM is transported as suspended matter or is not caught by the sediment basins, particularly during high-flow events. Sediment basins do not capture all the OM eroded from uplands and are biased toward the capture of larger particulate OM (Verstraeten & Poesen, 2000) over OM associated with fine particles. It is important to note that long-term (over thousands to millions of years) erosion rates in the Sierra Nevada (Milodowski et al., 2015; Riebe et al., 2001) and elsewhere in the western United States (Kirchner et al., 2001; Reneau & Dietrich, 1991) suggest that either these sediment basins underestimate erosion long-term rates or contemporary rates are much slower than long-term rates. At the catchments in this study, the sediment export based on sediment trapping ranged from 1.9 to $79.5 \text{ kg}\cdot\text{ha}^{-1}\cdot\text{yr}^{-1}$ from 2005 to 2011 (Stacy et al., 2015), but long-term denudation in the Sierra based on cosmogenic nuclide concentrations and geochemical mass balance is estimated to be 230 to $7,550 \text{ kg}\cdot\text{ha}^{-1}\cdot\text{yr}^{-1}$ (Riebe et al., 2004). While in-stream sediment transport would not match hillslope denudation rates due to the deposition of sediment in upslope depositional areas, one could presume that a large portion of sediment is eventually exported from these small catchments, perhaps to be deposited in an adjacent less steep catchment or in a lake or reservoir downstream (Meade et al., 1990). The most important effect of climate on Sierra Nevada denudation rates is likely to be in the short time scales (Riebe et al.,

2001). One explanation is that episodic events not captured in short-term assessments result in a substantial portion of sediment transport on longer time scales. Episodic high flows yielded most of sediment transport of both inorganic and organic material out of montane catchments in other Mediterranean environments (Farnsworth & Milliman, 2003; Gray et al., 2015; Rodríguez-Blanco et al., 2013).

Although the soil fractionation procedure used in this study is very useful in separating soil OM into ecologically meaningful fractions, it is also very time consuming, limiting the number of samples and replicates that can be analyzed. We present a comparison of soils and sediments but processed too few samples to assess statistical significance. We also did not determine if there was significant decomposition of eroded OM during transport or after deposition in the basins, or if there was change in aggregation or formation of organomineral associations prior to our analysis. Aggregates may be affected according to size, based on the greater relative importance of *MaA-C*, and concomitant drop or no change in *MiA-C*, in sediments compared to crest and backslope soils. Those differences may indicate that aggregates are reconfigured during or soon after erosional redistribution or that larger aggregates are more efficiently captured in the basin, while smaller aggregates are transported farther (Nadeu et al., 2011; Verstraeten & Poesen, 2000). Nevertheless, we argue that any aggregate modifications would occur in natural storage locations and the collected sediment represents changes that would occur within the stream or after deposition further down the stream channel.

5. Conclusions

We found that stabilization mechanisms protecting OM vary by year, but these interannual shifts in stabilization mechanisms are overwhelmed by the interannual changes in OM concentration due to changes in stream discharge. Interannual precipitation differences were more important for overall OM transport than the difference in elevation (i.e., precipitation form). Most sediment OM captured during export from these granitic clay-poor catchments was largely free OP or was contained in *MaA* and is not as protected from decomposition as mineral associated OM. Compared to the mineral soil from contrasting landform positions within each catchment, the sediment was enriched in free OP and had coarser mineral particles; the prevalence of coarser mineral fractions likely decreases the opportunity for protection through organomineral associations in these sediments. These changes from soil to sediment are due to selective erosion processes that mobilize organic material from the surface and to the selective deposition of denser and larger particles in the settling basins. Based on the two water years examined, we conclude that higher precipitation would likely result in greater transport of denser particles and possibly the mobilization of stabilized, mineral-associated C from deeper soil layers. The fraction of mineralizable C increases in eroded sediments but decreases in high stream discharge years as the sediment becomes more OM-poor. Unless buried downstream or otherwise stabilized, sediment OM is more likely to decompose than soil OM, thus decreasing the chance for erosion to result in a net C sink in these headwater montane catchments.

Acknowledgments

This study was conducted in the Kings River Experimental Watersheds (KREW) project, which was established and is managed by the Pacific Southwest Research Station of the USDA Forest Service. The KREW study was implemented using funds from the National Fire Plan of the USDA Forest Service. Additional funding for this work was provided by the Pacific Southwest Research Station of the Forest Service, the Southern Sierra Critical Zone Observatory Project of the National Science Foundation (EAR-0725097, 1239521, and 1331939), an NSF award to A. A. Berhe, S. C. Hart, and D. W. Johnson (EAR-1147977), and a Graduate Research Council grant from the University of California, Merced. We thank Michael Kaiser, Matthew McClintock, and members of the Berhe and Hart laboratories at UC Merced, researchers from the Southern Sierra Critical Zone Observatory, and staff from the Pacific Southwest Research Station for assistance in the field and laboratory. Data are available from the Southern Sierra Critical Zone Observatory digital library (https://eng.ucmerced.edu/snsjho/files/MHWG/Field/Southern_Sierra_CZO_KREW/) or from criticalzone.org/sierra/data.

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